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Significance of spatial variability in precipitation for process-oriented modelling: results from two nested catchments using radar and ground station data

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Abstract. The importance of considering the spatial distribution of rainfall for process-oriented hydrological modelling is well-known. However, the application of rainfall radar data to provide such detailed spatial resolution is still under debate. In this study the process-oriented TAC^D (Tracer Aided Catchment model, Distributed) model had been used to investigate the effects of different spatially distributed rainfall input on simulated discharge and runoff components on an event base. TAC^D is fully distributed ($50\times50\,\mathrm{m}^2$ raster cells) and was applied on an hourly base. As model input rainfall data from up to 7 ground stations and high resolution rainfall radar data from operational C-band radar were used. For seven rainfall events the discharge simulations were investigated in further detail for the mountainous Brugga catchment ($40\,\mathrm{km}^2$) and the St. Wilhelmer Talbach ($15.2\,\mathrm{km}^2$) sub-basin, which are located in the Southern Black Forest Mountains, south-west Germany. The significance of spatial variable precipitation data was clearly demonstrated. Dependent on event characteristics, localized rain cells were occasionally poorly captured even by a dense ground station network, and this resulted in inadequate model results. For such events, radar data can provide better input data. However, an extensive data adjustment using ground station data is required. For this purpose a method was developed that considers the temporal variability in rainfall intensity in high temporal resolution in combination with the total rainfall amount of both data sets. The use of the distributed catchment model allowed further insights into spatially variable impacts of different rainfall estimates. Impacts for discharge predictions are the largest in areas that are dominated by the production of fast runoff components. The improvements for distributed runoff simulation using high resolution rainfall radar input data are strongly dependent on the investigated scale, the event characteristics and the existing monitoring network.

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1 Introduction

The spatial variability of rainfall is often identified as the major source of error in investigations of rainfall-runoff processes and hydrological modelling (O'Loughlin et al., 1996; Syed et al., 2003). Especially for smaller catchments and for runoff processes that respond directly to precipitation detailed rainfall information is necessary (Woods et al., 2000). However, the spatial variability of precipitation can be very strong. Mean diameters of rain cells vary hugely for different climates and event types and hence, "typical" values are region dependent. For example, diameters have been estimated between 15 km (Luyckx et al., 1998) and one to 5 km (Woods et al., 2000) or an area of 1–2 km^2 (Thomas et al., 2003), and such cells can move significantly during events. Obviously, such detailed information on rainfall distribution and heterogeneity is unobtainable with a standard ground station density of 1 station per 20 km^2 (Michaud and Sooroshian, 1994).

In addition to errors in catchment precipitation, relatively small differences in catchment precipitation based on different rainfall input data might result in comparable large errors in simulated runoff (Sun et al., 2000) due to the spatial aggregation of rainfall information (Faures et al., 1995; Winchell et al., 1998). Some studies have found an increase of simulated runoff volumes (Michaud and Sorooshian 1994; Winchell et al., 1998) using spatial high resolution rainfall input data, while one study found a decrease (Faures et al., 1995). Partly, catchment runoff responded more sen-

Table 1. Basin characteristics of the Brugga basin and the subbasin St. Wilhelmer Talbach.

	Basin properties	
Name	Brugga	St. Wilhelmer Talbach
Elevation range (m)	438–1493	633–1493
Area (km^2)	40	15.2
Geology	Gneiss covered by drift	Gneiss covered by drift
Dominant vegetation type	Forest and pasture land	Forest and pasture land
% forested	71	73.4
Mean precipitation (mm)	1750	1853
Mean runoff (mm)	1195	1301
Mean evapotranspiration (mm)	555	552

sitively to temporal than to spatial resolution of precipitation data (Krajewski et al., 1991). Conversely, Obled et al. (1994) have found no significant improvement in hydrological predictions using temporally higher distributed rainfall in a medium-sized rural catchment, although they emphasised the possibility of contradictory results for smaller urbanized or larger rural catchments.

The spatial and temporal distribution of precipitation can have different relevance for distinct runoff generation processes. Modelled Hortonian overland flow is likely to be more sensitive to spatially and temporally averaging of precipitation than saturation excess runoff, increasing with a more spatially distributed rainfall input (Michaud and Sorooshian, 1994; Winchell et al., 1998). Furthermore, different spatio-temporal variable characteristics of rain cells, e.g. storm cell position or volume of the storm core, cause different impacts to runoff generation mechanism depending on catchment and event characteristics (Syed et al., 2003). In addition to runoff volume and peak flow, also the timing of hydrological response is influenced by spatial distribution of rainfall input (Krajewski et al., 1991; Ogden et al., 2000). Sun et al. (2000) improved the timing of peak flow estimations using higher distributed rainfall data. However, improvements of flow predictions depend on a wide range of factors such as investigated catchment scale, rainfall and catchment characteristics, runoff generation mechanism and applied model (Ogden et al., 2000; Arnaud et al., 2002).

Rainfall radar data provide the opportunity to apply spatially distributed rainfall data in distributed catchment modelling (e.g. Uhlenbrook and Tetzlaff, 2005). Especially in catchments with coarse raingauge networks, radar data can contribute to process-realistic distributed runoff simulations (Michaud and Sorooshian, 1994; Lange et al., 1999; Woods et al., 2000). Although in recent years rainfall radar data have been utilized more and more in hydrological studies, the benefit of radar data is still controversial. For example, a number of studies exists which focus on descriptions of rain drop size distribution, variability in Vertical Profile Reflectivity (VPR) or other influencing factors if transferring measured reflectivities in rainfall intensities (Smith and Krajewski, 1993; Fabry, 1997; Borga et al., 1997; Hirayama et al., 1997; Uijlenhoet and Sticker, 1999; Grecu and Krajewski, 2000a, b; Borga, 2002). These authors developed techniques for an improved estimation of rainfall rates from radar reflectivities for hydrological application and thus, an improvement of runoff modelling, although they acknowledge that significant uncertainties remain. A relatively large uncertainty, which is associated with rainfall intensities estimated from reflectivities, affects mainly the rainfall peaks (Morin et al., 2001). In most cases, operational available data are not sufficient enough regarding their quality due to the single-polarization measurement. Only few studies exist, which apply approaches with an adequate effort in correction of the radar data (Winchell et al., 1998; Ogden et al., 2000; Carpenter et al., 2001).

This study seeks to add to current research regarding methods of defining spatial variability in precipitation and the potential value of radar data. It has three specific aims: Firstly, to develop a methodology for the adjustment of the operational available radar data for single events for subsequent hydrological model applications. Secondly, to investigate influences of different rainfall data sources on estimated catchment precipitation. Thirdly, to examine effects of different spatially distributed rainfall inputs on simulated runoff at the event scale in two nested catchments. To explore these questions, two nested, meso-scale catchments in the Southern Black Forest Mountains, Germany, were investigated. The catchments are characterized by distinct patterns of direct runoff producing areas. They are equipped with a dense rainfall station network and one weather radar station.

2 Study size

The study was performed in the meso-scale Brugga catchment ($40\,km^2$) and its subcatchment St. Wilhelmer Talbach ($15.2\,km^2$) located in the Southern Black Forest Mountains, southwest Germany (Fig. 1, Table 1). The Brugga basin is a pre-alpine mountainous catchment with a mean elevation of

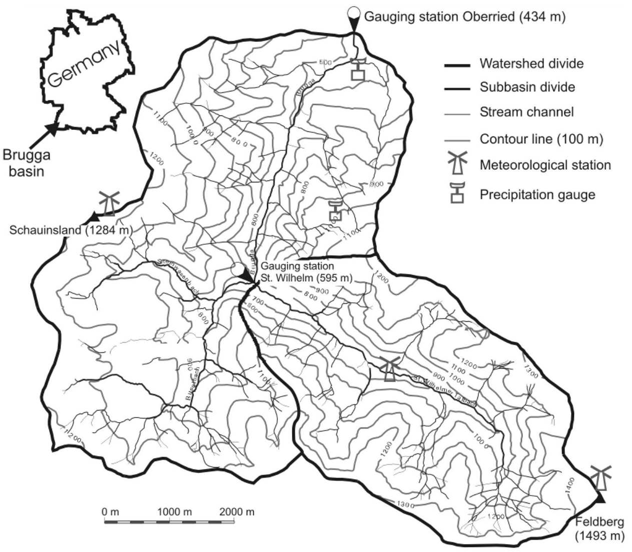

Fig. 1. The investigated catchments Brugga and St. Wilhelmer Talbach and its instrumentation network.

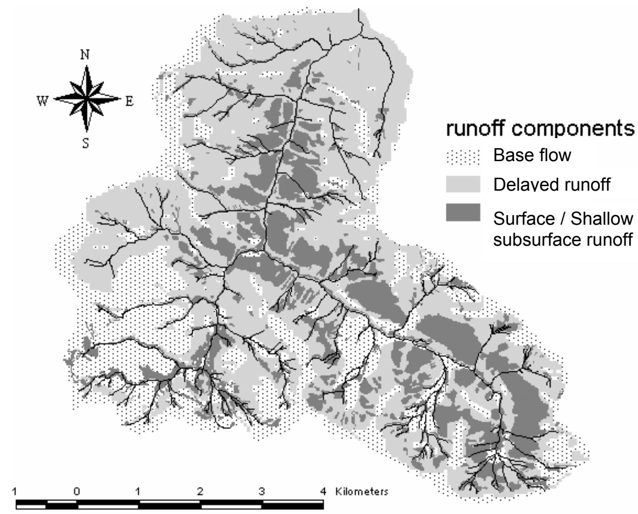

Fig. 2. Simplified spatial distribution of dominant runoff generation areas: base flow, delayed runoff (interflow), surface and shallow subsurface runoff (fast runoff).

about 986 m a.s.l.. Steep hillslopes, bedrock outcrops, deeply incised and narrow valleys, and gentler areas at the mountaintops characterize the mountainous part of the basin. The gneiss bedrock is covered by brown soils, debris and drift of varying depths at the hillslopes (0–10 m). Soil hydraulic conductivity is generally high: the infiltration capacity is too high to generate infiltration excess except small settlements with sealed surfaces.

The morphology is characterised by moderate to steep slopes (75% of the area), hilly hilltops and uplands (about 20%), and narrow valley floors (less than 5%). The overall average slope is 19°, calculated with a $50\times50\,\mathrm{m}^2$ digital elevation model.

Mean annual precipitation is 1750 mm; mean annual runoff is 1195 mm. Between both catchments, specific mean daily flow is comparable with $39.1\,\mathrm{l\,s^{-1}\,km^{-2}}$ (Brugga) and $41.3\,\mathrm{l\,s^{-1}\,km^{-2}}$ (St. Wilhelmer Talbach) (Table 2), but maximum flows vary with maximum recorded flows of $840\,\mathrm{l\,s^{-1}\,km^{-2}}$ (Brugga) and $763\,\mathrm{l\,s^{-1}\,km^{-2}}$ (St. Wilhelmer Talbach) (Table 1). Due to the strong variability of elevation, slope and exposure caused by the deeply incised valleys the catchment is characterised by a large heterogeneity of all climate elements, in particular precipitation. This causes spatially and temporally irregular elevation-precipitation gradients within the basin and articulated luv-lee i.e. rain shadow effects.

Experimental investigations using artificial and natural tracers showed the importance of three main flow systems (Uhlenbrook et al., 2002, 2005): (i) fast runoff components (surface and shallow subsurface runoff) which are generated on sealed or saturated areas or, additionally, on steep highly permeable slopes covered by boulder trains; (ii) slow base flow components (deep groundwater) are connected with fractured rock aquifers and the deeper parts of the weathering zone, and (iii) an intermediate flow system originates mainly from (peri-) glacial deposits of the slopes (shallow ground water). These are mainly delayed runoff components compared to the surface and near-surface runoffs. However, they can also contribute to flood formation depending on the antecedent moisture content. A simplified spatial delineation of hydrological homogeneous regions – generating the three main runoff components base flow, interflow as well as surface and near surface runoff – is shown in Fig. 2. Most parts of the study catchment are covered by glacial and periglacial drift cover and hence, influenced by interflow processes. The extent of areas generating mainly fast runoff components is defined by saturated and sealed areas as well as very steep hillslopes (>25°).

3 Data and methods

3.1 Precipitation data

3.1.1 Ground station data

For radar data calibration, up to 11 ground stations were – event dependent – available within and nearby the catchment boundaries. Nine of these ground stations are located in a circumference of maximal 30 km of the investigated catchments at elevations between 200 and 1010 m a.s.l.. Ground stations measured with a time resolution of 1–15 min. More ground stations within the catchments were available but they were measuring on a coarser time resolution and hence, were not used for radar data calibration.

For the subsequent runoff simulations, up to seven ground stations were used, located within (five stations, see Fig. 1) or

Table 2. Specific discharge values for the investigated catchments (data source: LfU 1999).

	Brugga 40 km^2	St. Wilhelmer Talbach 15.2 km^2
Period	1934–1998	1954–1997
Highest recorded flow (l s^{-1} km^{-2})	840	763
Mean high flow (l s^{-1} km^{-2})	342	406
Mean daily flow (l s^{-1} km^{-2})	39.1	41.3
Mean low flow (l s^{-1} km^{-2})	9.03	7.9
Lowest recorded flow (l s^{-1} km^{-2})	2.5	1.3

very close (two stations, not displayed) to the Brugga basin. Basin precipitation was interpolated using an 80:20 combination of the Inverse Distance Weighting (IDW) method (80%) and an elevation gradient (20%). This was done because of an observed elevation dependence of precipitation that was found for longer time intervals (monthly, yearly), but which was not always observed for shorter time steps. During storms the location of the rain cell is more important than elevation, which explains the unequal weighting of the two interpolation methods. Consequently, the used interpolation scheme is a compromise to capture the spatial distribution during shorter time intervals but also to reproduce the long term pattern. The precipitation measurement error caused by wind was corrected according to the approach of Schulla (1997) that differentiates between liquid and solid precipitation.

The IDW method is often used as an alternative to Kriging to compute the rainfall covariance function (Odgen et al., 2000). The IDW method calculates a weighted average precipitation for each raster cell with a weight of d^{-2}, while d is the distance between the gauging station and the respective raster cell. Only stations within a radius of 6 km for each raster cell were considered for the calculation. The elevation gradient is a non-linear function that considers the mean annual increase of precipitation with height (Uhlenbrook et al., 2004). This gradient was kept constant within the basin, but varied for every modelling time step.

3.1.2 Radar data and adjustment methods

Weather radars are not measuring the rainfall intensity itself but the radar reflectivity. Reflectivities are converted into rainfall rates using the Z-R-relation

$$Z = \alpha * R^{\beta} <==> R = (Z/\alpha)^{1/\beta} = (10^{dBZ/10}/\alpha)^{1/\beta} \quad (1)$$

with

$$dBZ = 10 \log Z\,, \quad (2)$$

where Z is the reflectivity (mm^6 m^{-3}) and R the rain intensity (mm h^{-1}). α and β are fitting parameters.

The calculation of intensities from the measured reflectivities is influenced by numerous factors and includes high uncertainties (Uijlenhoet and Stricker, 1999). Reflectivities are strongly dependent on size of the raindrops, their density, rainfall type and characteristics. Therefore, different Z-R-relations arise according to seasonal and meteorological conditions (e.g. Smith and Krajewski, 1993; Quirmbach et al., 1999; Haase and Crewell, 2000).

The rainfall radar data used in this study are measured by a C-band Doppler radar with a wavelength of 3.75–7.5 cm and one elevation angle (0.5°). The rainfall radar station is located near the highest point of the Brugga catchment at the peak of the Feldberg Mountain (Fig. 1). The radar product is a quantitative DX product provided by the German Weather Service (DWD). The spatial resolution is 1 km×1° azimuth angle with a temporal resolution of 5 min. The data from 1998 have only dBZ classes with 4-dBZ steps due to a systematic measuring error during this time period. These technical problems were solved in 1999 and from then the resolution of dBZ values is 0.5.

The radar data were corrected for clutter by the German Weather Service using clutter maps. These clutter maps are compiled during a period when no precipitation echoes are relevant. Neither distance nor vertical reflectivity profiles corrections were conducted. A detailed description of the used DX product can be found at DWD (1997). Problems connected with these operational radar products available in Germany are discussed e.g. in Quirmbach (2003).

In general, for the correction of radar data two main basic approaches exist. Firstly, the correction of vertical profiles of reflectivities using different radar beam elevation angles (e.g. Andrieu et al., 1997; Creutin et al., 1997; Borga, 2002). The radar data used in this study were measured only with one elevation angle. Therefore, this approach could not be applied. Additionally, it can be assumed that – especially during convective events – small variabilities of reflectivities occur until a height where the 0°C isotherm is reached (Fabry, 1997). In summer, this border lies some kilometres above ground. Furthermore, variations of reflectivities are small near the radar site (Andrieu and Creutin, 1995). Both aspects, that radar data of convective events were used and for a study catch-

Table 3. Rain event characteristics.

Event	Date	No. of ground stations used for radar calibration	Max. radar reflectivity (dBZ)	Duration of precipitation event (h)	Total rain amount at ground station St. Wilhelm (mm)	α (−)	β (−)
1	27 July 1998	9	52	17	22	40	1.73
2	22 August 1998	9	36	15	33.8	50	1.12
3	4 September 1998	9	40	20	52.4	71	1.13
4	23 May 2002	11	47	15.75	17.9	52	2.16
5	25 May 2002	11	44	7.75	10.3	36	4.18
6	4 June 2002	10	50	1.75	21.2	40	1.66
7	6 June 2002	10	50	23.5	65.6	10	2.28

ment close to the radar site, let the authors assume that a vertical correction of the reflectivity profiles can be neglected in this case study.

Therefore, the second approach of adjusting radar-derived precipitation using gauge data was applied. The aim of this approach is to correct the estimated radar precipitation with gauge measurements (e.g. Adamowski and Muir, 1989; Seo et al., 1999; Sun et al., 2000; Vallabhaneni et al., 2002). A main error source in such radar data calibration is due to the drawback on appropriate ground station data, e.g. due to the lack of an adequate number of ground stations (Ciach and Krajewski, 1999). Ground station data can capture the temporal distribution of rainfall very well, but the spatial representation is often limited, especially in heterogeneous catchments with sparse ground station network. In contrast, radar data allow very detailed information about the spatial distribution of precipitation, but measurements have practical limitations in estimating rainfall totals.

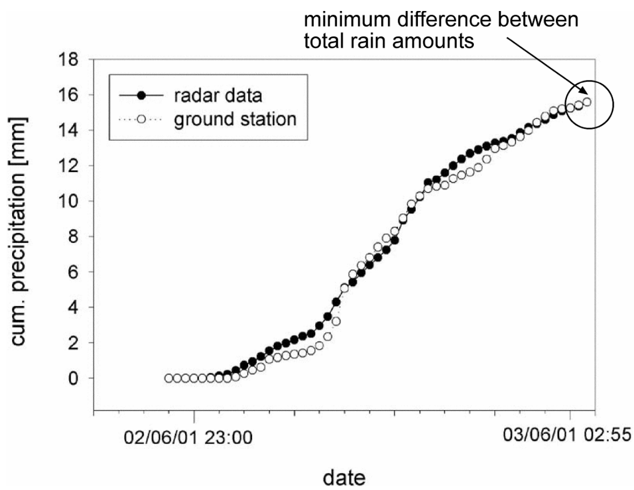

Fig. 3. Radar data calibration using the minimum square distance method for the cumulative curves of both rainfall data sets, showing minimized difference between the total precipitation amounts of both data sets as an additional constraint.

3.1.3 Radar data calibration at the event scale

Within this study, radar data were calibrated using the certain radar cell corresponding to the ground station data. Firstly, equal time intervals of 5 min between the radar and ground data were constructed for comparability of both data sets. Ground station data were either aggregated (sum to 5 min) or disaggregated. It became clear that an event and station dependent time shift correction between both data sets was necessary. Results showed that between both data sets a station and event dependent time shift correction of 5 to 15 min was necessary. Because of wind drift of falling precipitation a neighbouring pixel can be more representative than the direct corresponding pixel. Thus, an average of nine cells, i.e. the cell with the location of the rain gauge and all eight surrounding cells, was used to compare with the rain gauge measurements. Depending on event and station, a coefficient of correlation (r) between both data sets of more than 0.69 was obtained after time shift correction. Additionally, a visual check was executed to identify errors in the radar images e.g. ground clutter.

Afterwards radar data were adjusted with an automated algorithm using a tool for optimization and equation solving. By minimising the total square deviation between the cumulative precipitation curves of both data sets, the distribution of rainfall intensities in each time step is considered (Fig. 3). An additional constraint was to minimize the difference between the total precipitation amounts of both data sets. An optimum parameter set of α and β of the Z/R-relation for each event was then determined by automatically minimising both square deviation and differences of total rain amounts of all available ground stations. Optimum, but physically reasonable α and β parameters were determined. This non-linear adjustment avoids weighting higher rain intensities more significantly than lower rain intensities. Resulting Z/R-relations differ strongly between the single events (Table 3). In a next step, the measured radar reflectivities were transformed into rainfall intensities using constant in space event dependent Z/R-relations. Using these Z/R-relations the radar rainfall intensities were calculated for the whole catchment in a spatial resolution of 1 km×1° azimuth angle and a temporal resolution of 5 min using Arc Info GIS routines.

3.2 Hydrological modelling

In recent years, several hydrological models have been used at the Brugga basin and sub-basins (e.g. PRMS/MMS, Mehlhorn and Leibundgut, 1999; TOPMODEL, Güntner et al., 1999; HBV, Uhlenbrook et al., 1999). The application of these models and the results of the experimental studies led to the development of the TAC model, the Tracer Aided Catchment model (Uhlenbrook and Leibundgut, 2002). The aim was to develop a better process-realistic model to compute the water balance on a daily mode. TAC is a process-oriented, semi-distributed catchment model, which requires a spatial delineation of units with the same dominating runoff generation processes (cf. hydrotopes or hydrological response units).

The TAC model was advanced to the TAC^D model (TAC, distributed), a fully distributed raster model (Uhlenbrook et al., 2004). The spatial division was undertaken by delineating the catchment into units sharing the same dominating runoff generation processes. The units were converted into 50×50 m^2 raster cells that are connected by a single flow algorithm. Channel routing is modelled with a kinematic wave approach (implicit, non-linear). The whole model is integrated into the GIS PC-Raster (Karssenberg et al., 2001).

The TAC^D model was applied to the Brugga basin using the period 1 August 1995–31 July 1996 for model calibration (further details are given in Uhlenbrook et al., 2004). It was initialised over a period of three months using estimated values for the hydrological storages prior to this period. The calibrated parameter set was used for modelling the St. Wilhelmer Talbach sub-basin without re-calibration. To evaluate model goodness the model efficiency $R_{eff}(Q)$ (–) (Nash and Sutcliffe, 1970) and the model efficiency using logarithmic runoff values $R_{eff}(\log Q)$ (–) were used. Good simulation results were obtained at Brugga catchment for the model calibration period ($R_{eff}(Q)$=0.94; $R_{eff}(\log Q)$=0.99) and validation period (three years record; $R_{eff}(Q)$=0.80; $R_{eff}(\log Q)$=0.83) after a split-sample test. A multiple-response validation using additional data, including tracer data, demonstrated the process-realistic basis of the model with its simulated runoff components (Uhlenbrook et al., 2004).

The calibrated radar data with a temporal resolution of 5 min were aggregated to 1 h intervals to serve as input for the TAC^D model. The original spatial resolution of the polar co-ordinate grid of 1 km×1° azimuth angle was disaggregated to a 50×50 m^2 grid using GIS ARCINFO grid module. Due to technical limitations of the radar measurement, a small area around the radar device needed to be “filled” with ground data measurements.

The following methodology was devised to compare the impact of the two precipitation inputs on event runoff simulations. The model was run twice, each time with the same initialisation period (eight months), parameter values (determined during model calibration) and input data sets, but with different basin precipitation maps for each time-step of the investigated events. This approach has the advantage that the model runs continuously and thus, the spatial and temporal variable soil moisture and groundwater storages are modelled reasonably prior the investigated event. This is a prerequisite for process-oriented modelling, which could not have been fulfilled if the events were modelled separately and independently from the previous hydrological conditions.

4 Results

4.1 Influence of different rainfall input data on the estimated catchment rainfall

The seven single rain events investigated varied in their measured maximum radar reflectivities of up to 52 dBZ (Table 3). Due to the contrasts in event characteristics, event 6 and 7 are mainly presented and discussed within this study. Event 6 is the most convective event with very short duration and high rainfall intensities. Event 7 shows the highest precipitation amount and resulted in the highest flow due to the long event duration.

Example percentage deviations between the total rain amounts at the respective ground station and the corresponding radar cell for events 6 and 7 show clearly that there was neither systematically under- nor overestimation of the precipitation amount (Table 4). Occasionally, at single stations high deviations occur, but at station 7, which is situated near the centre of the St. Wilhelmer Talbach sub-catchment, the deviations can be neglected (<10%).

To examine the influence of different rainfall input data for basin precipitation, mean, maximum and minimum precipitation values were compared (Table 5). It becomes clear that the maximum and minimum values were more extreme – i.e. higher and lower – using radar data than ground station data. The high maximum values using IDW-elevation method for event 7 were due to a high value at only one ground station (Feldberg), while all other ground stations recorded precipitation amounts between 60–70 mm during this event. Although maximum intensities were higher with radar data, in most cases mean catchment precipitation was higher using the IDW-elevation regression. The interpolation with IDW does not fully account for the variability in rainfall in between the rain gauges and therefore, produces smoother precipitation fields and higher values if high ground station values are interpolated to larger areas of the basin.

Table 4. Percentage deviation of the total rain amount: radar from ground station value (%).

Station	Event 6	Event 7
1	−8	+2
2	+25	−19
3	+33	+73
4	+127	−13
5	+83	−13
6	−46	−14
7	+7	+9
8	0	−4
9	−31	−4
10	−42	+30

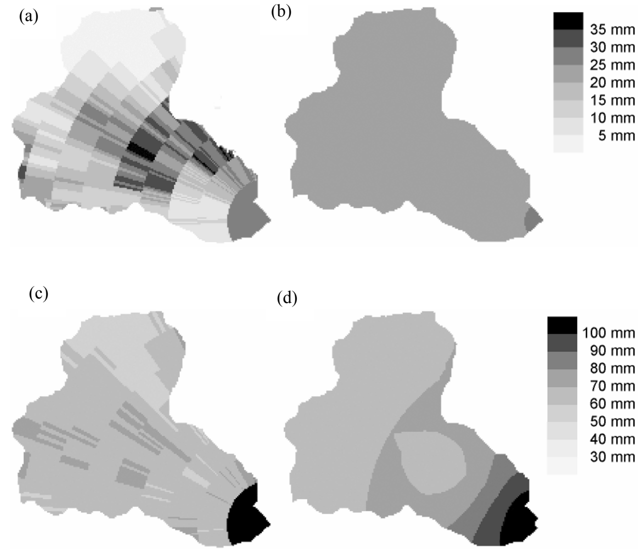

Fig. 4. Spatial distribution of basin precipitation (cumulated over whole event period) during the events 6 **(a)** radar, **(b)** ground station, and event 7 **(c)** radar and **(d)** ground station.

Using the different precipitation inputs resulted in large differences in the spatial delineation of the precipitation fields (Fig. 4). During the strong convective event 6 (duration: 1.75 h) the rain cell was mainly located in the St. Wilhelmer Talbach subcatchments (Figs. 4a and 4b), which is well represented by one ground station. The precipitation field with radar data was much more heterogeneous than with the IDW-elevation-regression method with precipitation ranges between 1 mm (minimum) and 38 mm (maximum) within the whole Brugga catchment. Due to the interpolation of rainfall, mean precipitation was 30% higher using the IDW-elevation-regression method than radar data, although maximum rainfall intensities were not captured using just ground station data.

Event 7 (Figs. 4c and 4d) was less convective, but with higher total rain amounts after a longer event duration (23.5 h). Again, maximum and minimum values (Table 5) were more extreme with radar data compared to application of ground station data. In addition, the precipitation field using radar data was more heterogeneous compared to the IDW-elevation-regression method, although differences in the total amounts were compensated because of the longer duration of the event. Again, higher total precipitation amounts were reached applying ground station data, which resulted in mean precipitation values 17% higher than using radar data.

4.2 Effects of different rainfall input on simulated discharge

Subsequently, the ground station data and the calibrated radar data were used as input for runoff simulation using TAC^D. For all investigated events model efficiency values (Table 6) were used for an assessment of the influence of different spatially distributed rainfall input on simulated runoff. In general, different spatial resolution as well as total event rainfall are likely to contribute to differences in the simulated runoff between the two rainfall data sources. Better simulation results – i.e. higher model efficiencies (Nash and Sutcliffe, 1970) – were gained in the smaller St. Wilhelmer Talbach catchment. It has to be noted that this catchment is relatively well covered by one ground station located near its centre. If Nash and Sutcliffe model efficiencies were used for interpretation as goodness-of-fit criterion, ground station application resulted more often in higher values due to the fact that the model was calibrated using ground station data. However, for some events (e.g. event 3) model efficiency values were unacceptable, regardless of which rainfall input was used. Using additional criteria for model goodness, in most cases the percentage deviation between simulated and observed peak discharge was less using ground station data. Neither type of input data resulted in a systematically under- or overestimation of peak discharge. For the St. Wilhelmer Talbach sub-catchment, results were less clear regarding one input resulting in better runoff simulations. In the Brugga catchment, also no clear pattern became clear that one rain input resulted in better simulation results than the other regarding discharge volume. However, discharge volumes in the Brugga catchment were more often overestimated, while in the St. Wilhelmer Talbach catchment they were more often underestimated.

Looking in further detail at the two contrasting events, it becomes clear that during event 6 the use of ground station data resulted in an overestimation of the simulated peak discharge of 52% compared with the observed hydrograph in the Brugga catchment (Fig. 5). Simulation with spatial higher resolution radar data resulted in an overestimation of only 17%. The discharge volumes were overestimated by 38% (ground station data) and 22% (radar data).

Table 5. Comparison of rainfall values at the respective $50\times50\,m^2$ raster cells in the Brugga catchment based on radar data and ground data using IDW-elevation regression method for interpolation (mm).

Event	Date	Radar			IDW elevation-regression		
		Mean	Min	Max	Mean	Min	Max
1	27 July 1998	22.8	14.5	38.5	25.9	15.8	32.5
2	22 August 1998	44.3	26	74.5	35.1	23.2	44.8
3	04 September 1998	41.1	16.5	78.5	39.1	26.9	51.1
4	23 May 2002	16.5	11	27	18.7	17.4	21.8
5	25 May 2002	8.3	4	17	11.2	10.1	14.2
6	4 June 2002	15.9	1	38	22.7	20.3	25.3
7	6 July 2002	60.5	0	80	72.2	64.0	110.2

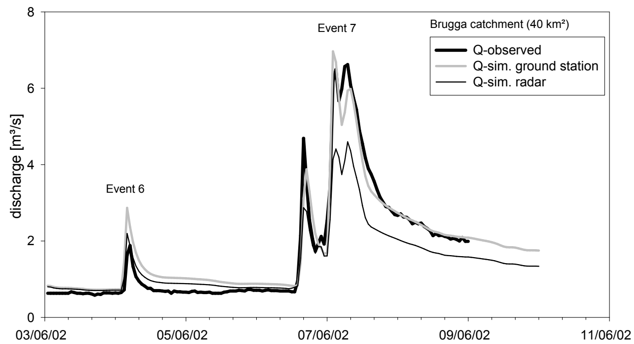

Fig. 5. Hydrographs of the events 6 and 7 for the Brugga catchment ($40\,km^2$).

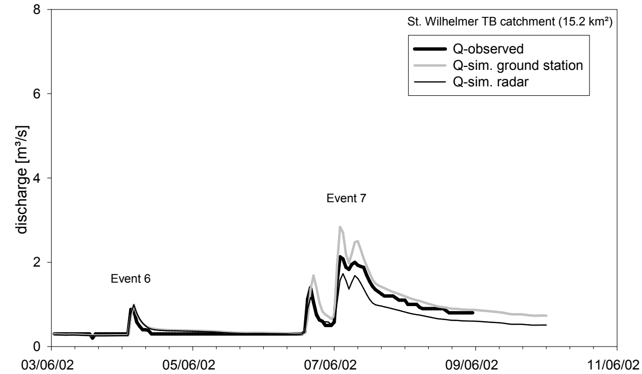

Fig. 6. Hydrographs of the events 6 and 7 for the St. Wilhelmer Talbach catchment ($15.2\,km^2$).

For interpretation of the hydrographs, it is important to consider the spatial distribution of precipitation in combination with the spatial delineation of the main hydrological response units (Fig. 2). The higher calculated catchment precipitation amount especially in the North of the Brugga catchment – due to the transformation of single ground station values for the whole sub-basin – resulted in this large overestimation in runoff simulation using ground station data. The effect was reinforced because this strong overestimation occurs in large parts of the subcatchment where fast runoff components are dominant (see Fig. 2). Model efficiencies for ground station data simulation were poor (R_{eff}=−0.99), but much better with radar data (R_{eff}=0.46). In this catchment, for which there are little ground station data, the use of radar data especially during such a highly localised event produced better runoff simulation results. If too high precipitation is determined in areas where fast runoff components are dominant, the errors in runoff simulation can be substantial.

The simulations in the St. Wilhelmer sub-catchment produced with both types of rainfall input data comparable results for event 6 (Fig. 6). Both rainfall data sets resulted in a slight peak and volume overestimation compared to the observed discharge, although no volume error occurred using ground station data. For peak discharge, deviations are less and also model efficiency values are higher using radar data, which can be explained again by a better capturing of precipitation characteristics for areas with fast runoff response.

During event 7 all model performance parameters were poorer using radar data as rainfall input compared to ground station data for the Brugga catchment. These simulation results were caused by an underestimation of the catchment precipitation during this event in this basin, although during calibration there was no systematic underestimation of the rain amount using radar data (Table 4). For this less localised event with the longer duration the main influencing factor for runoff simulation was the total difference between both rainfall data sets. Spatial distribution of rainfall in combination with runoff generation patterns is of less relevance. Thus, the simulated hydrograph using ground station data fitted much better with the observed hydrograph (Fig. 5).

For the St. Wilhelmer Talbach catchment model efficiency values for event 7 are good with $R_{eff}>0.8$ for both data sets. Peak discharge and volume are overestimated with ground station data (33% and 15%, respectively) but underestimated with radar data (−19% and −18%, respectively, Fig. 6).

Table 6. Statistical measures of goodness-of-fit for the runoff simulations based on radar data and ground station rainfall data for the two investigated catchments.

	Rain input	Brugga (40 km^2)	St. Wilhelmer Talbach (15.2 km^2)
Model efficiency (Nash and Sutcliffe, 1970) (−)			
Event 1	Ground station	0.75	0.55
	Radar	0.4	0.41
Event 2	Ground station	0.93	0.73
	Radar	0.42	0.61
Event 3	Ground station	0.01	0.84
	Radar	−0.88	−0.27
Event 4	Ground station	0.7	0.82
	Radar	0.64	0.76
Event 5	Ground station	0.53	0.57
	Radar	0.4	0.38
Event 6	Ground station	−0.99	0.59
	Radar	0.46	0.64
Event 7	Ground station	0.95	0.83
	Radar	0.71	0.82
Percentage deviation (simulated from observed peak discharge) (%)			
Event 1	Ground station	−14	−32
	Radar	−34	−34
Event 2	Ground station	−3	−34
	Radar	28	19
Event 3	Ground station	5	7
	Radar	21	41
Event 4	Ground station	−28	−11
	Radar	−32	−18
Event 5	Ground station	−24	−13
	Radar	−30	−17
Event 6	Ground station	52	13
	Radar	17	12
Event 7	Ground station	5	33
	Radar	−31	−19
Percentage deviation (simulated from observed discharge volume) (%)			
Event 1	Ground station	13	−15
	Radar	4	−13
Event 2	Ground station	16	−24
	Radar	54	20
Event 3	Ground station	86	19
	Radar	113	51
Event 4	Ground station	−7	−5
	Radar	10	−8
Event 5	Ground station	2	−2
	Radar	−4	−8
Event 6	Ground station	38	0
	Radar	22	6
Event 7	Ground station	0	15
	Radar	−24	−18

5 Discussion

The operational available radar data in Germany, which were used in this study, are only corrected for ground clutter by the provider. As such, no information about e.g. Vertical Reflectivity Profiles are available for those data. Thus, the efforts necessary for corrections using ground station data by the user are high (Quirmbach, 2003; Uhlenbrook and Tetzlaff, 2005) and the quality and the use of radar data for hydrological application is limited. The developed method is based on the adjustment of radar-derived precipitation using gauge data and considers the intensity distribution within the event in adjusting the cumulative curves of both data sets. As the intra-storm variability of rainfall intensity is considered explicitly using this approach, for a reasonable comparison with the radar data ground station data at high temporal resolution have to be applied. For radar calibration, not only ground stations within the catchment boundary but also those within a radius of not more than 20 km were used to extend the data set and to capture a wider spectrum of rainfall intensities. This method was developed for an event-based calibration. However, also for less convective rain-events and continuous hydrological modelling radar data can be calibrated using this methodology, because periods without rain do not have to be calibrated. Thus, calibration efforts can be minimized. For example, Terblanche et al. (2001) discuss limitations and shortcomings connected with observation and transformation of radar data and ongoing research to improve weather radar measurements.

The use of radar data resulted in higher maximum and lower minimum precipitation if the spatial distribution of the rainfall within the catchment was compared with ground data. Furthermore, the use of ground station data resulted in much smoother precipitation patterns due to the interpolation of point rainfall information to large areas. However, mean values of basin precipitation were in most cases higher using ground station data. In the larger catchment, shorter and more convective events lead to higher differences in catchment precipitation (i.e. total amount and spatial distribution) between both types of rainfall data. During such event conditions, it is more unlikely that localised rain cells are captured by the available ground station net. Such differences in either extreme values or total rain amounts are likely to have crucial effects for subsequent hydrological modelling (e.g. Michaud and Sorooshian, 1994). In addition, Syed et al. (2003) have found that the position of the storm core relative to the outlet becomes more important for runoff simulation with increasing catchment size.

Using spatially higher resolution rainfall data, some authors found an increase in runoff volume (e.g. Michaud and Sorooshian, 1994). However, Faures et al. (1995) emphasised a decrease. Within this study, 41% of the investigated cases resulted in an increase in runoff volume using radar data. In 53% of the cases, volumes were higher using ground station data, which are often less variable than radar data. Furthermore, deviations in peak discharge were less using ground station data. However, in this study two rainfall data types were compared and not only different spatial resolutions of one data type. Thus, errors might be caused already during data calibration.

Generally, for evaluations of the goodness of simulation results based on a given precipitation input, several model performance values should be used to capture the whole spectrum of effects, i.e. changes in peak, volume and timing. No clear patterns were obvious that one rainfall input resulted in better simulations than the other. For example, for the highly convective event (event 6) errors in runoff simulation were less if spatially high resolution radar data were applied. This was obvious by the much better model efficiency values and fewer deviations in both peak discharge and discharge volume for both catchments. Particularly in parts of the basin, which are characterised by fast runoff response, the correct detection of the rainfall pattern using highly distributed radar data was important. But in most investigated cases model efficiencies were poorer and percentage deviations were higher using radar data. Hence, the advantage of higher spatial resolution is likely to be overridden by limitations in data quality.

For single events with a longer duration, the spatial distribution of precipitation has less influence on mean catchment precipitation because differences in rainfall less variable. The differences in precipitation might be balanced or smoothed by the non-linear, antecedent and event dependent response in runoff generation processes, especially in mesoscale catchments (e.g. Grayson et al., 1997; Woods et al., 2000). Hence, differences in precipitation might not result in the same degree of differences in the simulated hydrographs. In smaller catchments differences in distribution of the precipitation seem to have a much larger influence on the runoff simulation even if in some small scales precipitation might not vary considerably in space anymore and hence, it is likely that precipitation has no impact on runoff generation. In general, the use of distributed, process-oriented models allows the use of detailed information and complex data sets, and the analysis of many details in hydrological predictions (Butts et al., 2004; Uhlenbrook et al., 2004). However, the effects of the detailed information for any runoff modelling system need to be understood and additional data set needs to be utilized adequately by the applied model. In such case, also the effects of different input data on many model outputs (e.g. the changing contribution of runoff components) can be analysed. The potential value of operational available radar data, which themselves involve many uncertainties, remains controversial (e.g. Georgakakos and Carpenter, 2003; Gourley and Vieux, 2003). Further research is needed regarding methods which define spatial variability in precipitation and hydrological response.

Within this study it was demonstrated clearly that rainfall overestimation could have substantial impact for the flood prediction especially if such overestimation occurs in areas that are dominated by the formation of fast runoff components. The results have shown that the combination of the advantages of both – the high spatial resolution of radar data and the high temporal of ground station data – is an important step towards increased consideration of variability in rainfall information. Consequently, the importance of the rainfall input data for flood prediction can be very large, and this should be considered as much as the nowadays frequently discussed parameter uncertainty (e.g. Beven and Binley, 1992; many papers since then) when using such process-oriented models. Both sources of uncertainties in combination with the model structure uncertainty (caused by the limited process understanding and process heterogeneity, e.g. Grayson et al., 1992; Seibert, 1999) sum up to the overall model prediction uncertainty (Sivapalan et al., 2003).

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